AIR-OCEAN SURFACE HEAT EXCHANGE (AOSHE) MODEL AND LOW FREQUENCY UNSTABLE MODES IN ATMOSPHERE AND OCEAN

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1. INTRODUCTION

Several important mechanisms of air-ocean interaction for the El Nino and Southern Oscillation (ENSO) phenomenon have been developed in the past decade: ocean wave propagation, delay-oscillator, two equilibrium states, and air-ocean coupled instabilities. Due to crude parameterization of thermodynamical processes in both atmosphere and ocean, these theories either cannot explain the transition between El Nino and La Nina (e.g., coupled instability theories), or need an artificially setting-up criterion to make such a transition (e.g., slowly propagating oceanic Rossby wave theory). The irregularity of ENSO implies that the ENSO events cannot be explained as a pure wave propagation. More detailed research on thermodynamics in both ocean and atmosphere is needed before we run the oceanic and atmospheric GCMs.

2. RELATIONSHIP BETWEEN PERTURBATIONS OF SST AND OML DEPTH

The thermodynamics of upper ocean, commonly used in the coupled models, can be summarized by (Hirst, 1986)

$$\frac{\partial T'_s}{\partial t} + u'_w \frac{\partial \overline{T}_s}{\partial x} = \kappa(\sigma h'_w - T'_s), \quad \frac{\partial \overline{T}_s}{\partial x} < 0 \tag{1}$$

where T'_s, u'_w, h'_w are perturbations of SST, zonal currents, and OML thickness. $\partial \overline{T}_s / \partial x$ is the mean zonal SST gradient. κ and σ are two parameters. If κ is large, Eq.(1) becomes

$$T'_s = \sigma h'_w \tag{2}$$

The expressions show that the increase of the OML thickness ($h'_w > 0$) leads to the increase of SST ($\partial T'_s / \partial t > 0$, or $T'_s > 0$). **Is this type of thermodynamics correct ? Not really.** It is correct only for a cooling OML, where the outgoing heat flux is greater than incoming heat flux. The thicker the layer, the less cooler of the layer. Thus, the relationship between perturbations of SST and OML thickness is given by

$$\frac{\partial T'_s}{\partial t} \sim h'_w + \cdots \tag{3a}$$

For a warming OML, on the contrary, the thinner the layer, the hotter the layer. Therefore, the relationship between perturbations of SST and OML thickness is written by

$$\frac{\partial T'_s}{\partial t} \sim -h'_w + \cdots \tag{3b}$$

The fact that only (3a) is chosen as a thermodynamical component for the ocean part in coupled air-ocean models makes the current ENSO theories quite incomplete.

3. AN OML SWITCHER

Arguments are cast in terms of simple mixed layer models (Chu et al., 1990; and Chu and Garwood, 1991), where it is assumed that the temperature and velocity are uniform over some depth, h_w, called OML depth, and that the penetration depth of solar radiation is much smaller than h_w. With these assumptions, one can write

$$\frac{\partial T_s}{\partial t} + \vec{u}_w \cdot \nabla T_s = \frac{Q_0 - \Lambda_w Q_{-h}}{\rho_{w0} c_{pw} h_w} \tag{4}$$

where T_s is the SST; $\vec{u}_w$ is the horizontal velocity; ρ_{w0} is the characteristic sea water density; c_{pw} is the sea water specific heat under constant pressure; Q_0 is the net surface heat flux, downward positive; Q_{-h} is the entrainment heat flux at the base of the OML computed by

$$\frac{Q_{-h}}{\rho_{w0} c_{pw}} = w_e (T_s - T_{-h}) \tag{5}$$

where w_e and T_{-h} are the entrainment velocity and the temperature at the base of the OML.

Entrainment velocity can also be parameterized in terms of OML TKE balance. If salinity effect is not considered at present, sources and sinks of TKE at ocean surface are wind work (proportional to wind speed cubed) and buoyant damping (or forcing) due to surface warming (or cooling).

$$w_e = \frac{(C_1 u_{w*}^3 - C_2 \alpha g h_w Q_0 / \rho_{w0} c_{pw})}{g h_w \alpha (T_s - T_{-h})} \tag{6}$$

where C_1 and C_2 are tuning coefficients, and u_{w*} is the water surface friction velocity, which is proportional to the air surface friction velocity u_{a*}.

Λ_w is the Heaviside function of $(C_1 u_{w*}^3 - C_2 \alpha g h_w Q_0 / \rho_{w0} c_{pw})$. Λ_w is an OML switcher: whose value equals 1 for the entrainment regime, usually associated with strong surface wind forcing; and equals 0 for the shallowing regime (the OML thickness taken the Monin-Obukhov length scale), usually associated with weak surface wind forcing.

4. AIR-OCEAN FEEDBACK MECHANISM

The tropical ocean surface generally receives heat, i.e., $Q_0 > 0$. The tropical subcloud layer is usually dominated by the mean easterlies, and the atmospheric deep

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convection is usually located at the upward branch of the zonal circulation. In the east (or west) of the deep convective region, the surface wind speed is enhanced (or reduced), and therefore the OML depth is increased (or decreased). Under weak wind forcing, the decrease of OML depth ($h'_w < 0$) leads to the augmentation of SST perturbation ($\partial T'/\partial t > 0$) and in turn to a production of the atmospheric convection to the west of the deep-convective region (Fig.1a). However, under strong wind forcing, the increase of OML depth ($h'_w > 0$) leads to the augmentation of SST perturbation ($\partial T'/\partial t > 0$) and in turn to a production of the atmospheric convection to the east of the deep-convective region (Fig.1b).

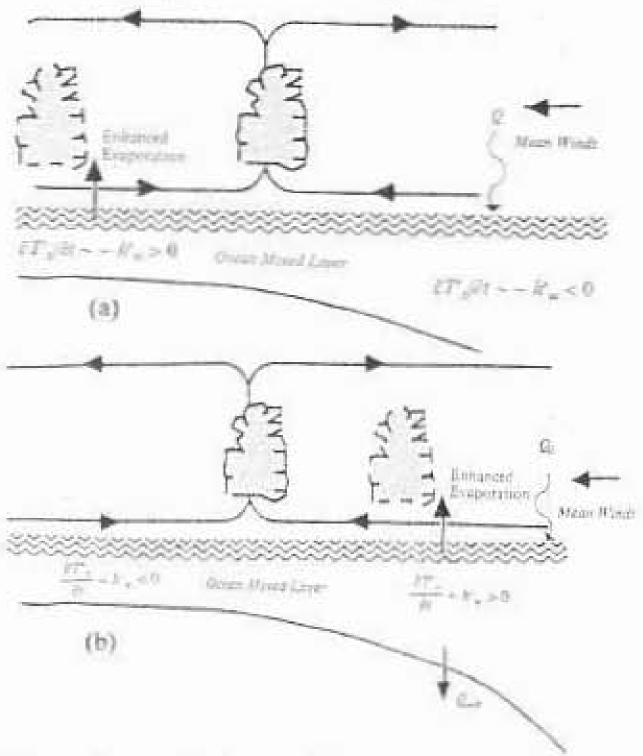


Fig. 1. Atmospheric onvection generated by air-ocean feedback under (a) weak wind forcing, and (b) the strong wind forcing.

5. AOSHE MODEL

The AOSHE model (Fig.2) is a coupled system consisting of the Wind-Induced Surface Heat Exchange (WISHE) model (Yano and Emanuel, 1991) and the Ocean Mixed Layer (OML) model (Chu and Garwood, 1991). Above the air-ocean interface, it reduces to the ordinary WISHE model (Yano and Emanuel, 1991) if the SST perturbation vanishes. Below the air-ocean interface, it is the ordinary OML model (Chu and Garwood, 1991). The AOSHE model is solved analytically as an eigenvalue problem (Chu, 1991ab) for two different surface conditions: strong and weak surface wind forcing. For strong wind forcing, there are two important parameters: γ ~ air-ocean surface coupling coefficient, and γ_e ~ entrainment parameter. For weak wind forcing, there is only one parameter, γ. When the surface wind stress is strong,

$$C_1(C_D\rho_{a0}/\rho_{w0})^{3/2}(u_b^2+v_b^2)^{3/2} - C_2\alpha g h_w Q_0/\rho_{w0}c_{pw} > 0,$$

the AOSHE model predicts the generation of unstable eastward propagating mode with phase speed (0.5-1.5 m/s) and the maximum growth rate (1.8/yr) appearing at 3-4 zonal wavenumbers (Fig.3), where the reasonable values of the model parameters for the entrainment regime of OML are: $\gamma = 0.02$, $\gamma_e = 0.75$.

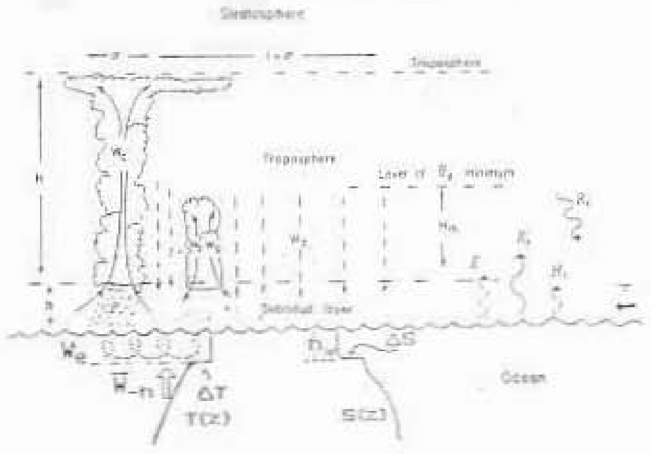


Fig. 2. Schematic presentation of the AOSHE model.

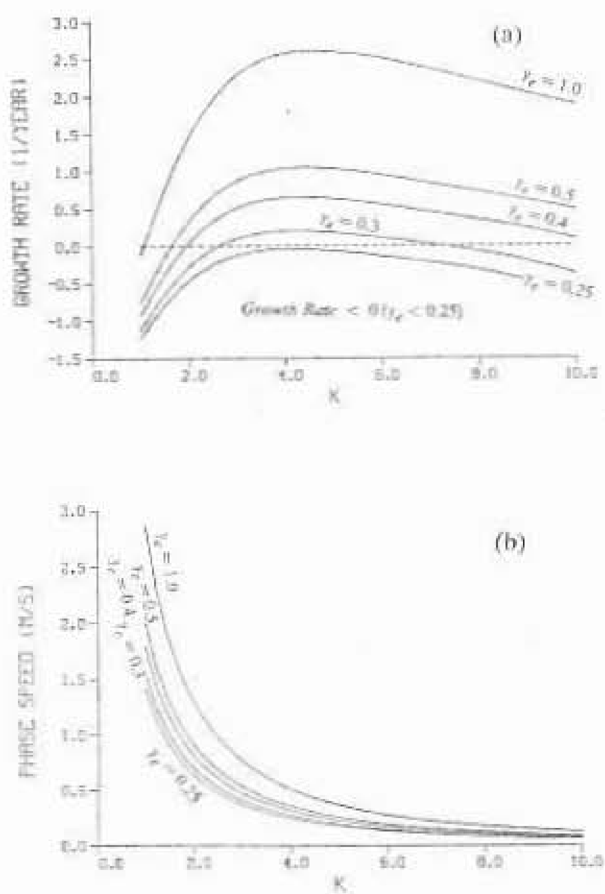


Fig. 3. Dependence of the AOSHE mode (under strong wind forcing) on the parameter γ_e: (a) growth rate (yr^{-1}), and (b) phase speed (ms^{-1}) for $\gamma = 0.02$.

When the ocean surface is under weak surface wind forcing (equivalent to strong surface warming),

$$C_1(C_D\rho_{a0}/\rho_{w0})^{3/2}(u_b^2+v_b^2)^{3/2} - C_2\alpha g h_w Q_0/\rho_{w0}c_{pw} < 0,$$

the AOSHE model predicts the generation of unstable westward propagating mode with phase speed (0.4 m/s) and the maximum growth rate (3.5/yr) appearing at the lowest zonal wavenumber (Fig.4), where the reasonable value of the model parameters for the shallowing regime of OML is: $\gamma = 0.03$.

This implies that there is an inherent switcher in the OML. Shifting the OML from one regime to the other regime only depends on the model variables: u_b, v_b, h_w, etc.

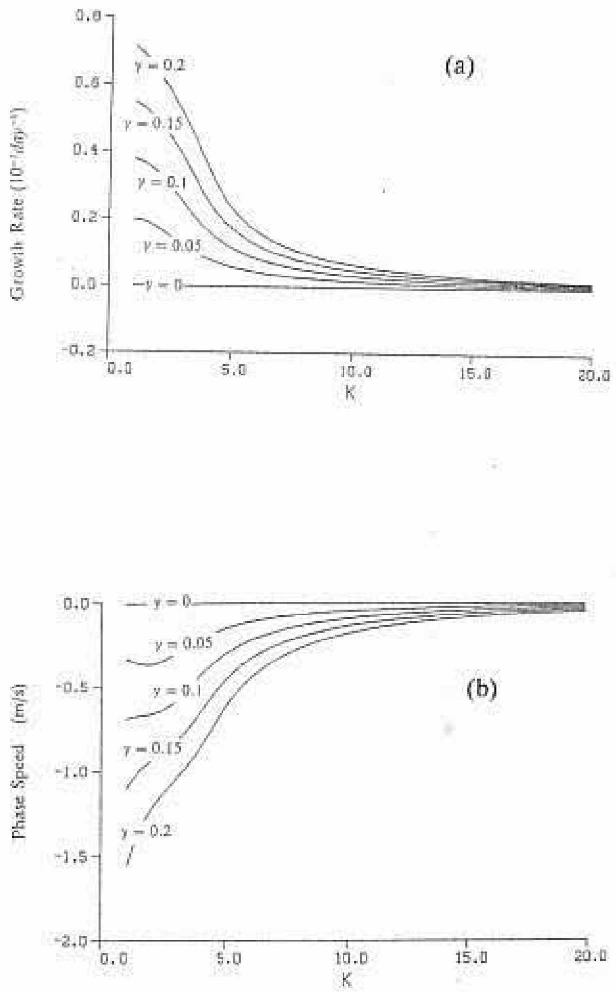


Fig. 4. Dependence of the AOSHE mode (under weak wind forcing) on the parameter γ: (a) growth rate and (b) phase speed (ms^{-1}).

6. EL NINO/LA NINA CYCLE

Based on the AOSHE model results, I offer a new theory for the La Nina/El Nino transition depicted as follows.

(a) La Nina Conditions Prevailed Stage

Starting from the typical La Nina conditions, atmospheric deep convection (denoted by "Convection A") is developed in the western Pacific (Fig.5a). The surface winds in the central and eastern Pacific are enhanced due to the same direction of the mean easterlies and the easterlies associated with the zonal circulation induced by Convection A. The OML is under strong surface wind forcing. The unstable convective disturbances (denoted by "Convection B") are generated in the east of Convection A and propagating eastward (Fig.5b) with phase speeds 0.5-1.5 m/s, and maximum growth rate (~ 1.8 /year) appearing at the 3-4 zonal wavenumber.

(b) Mature of La Nina

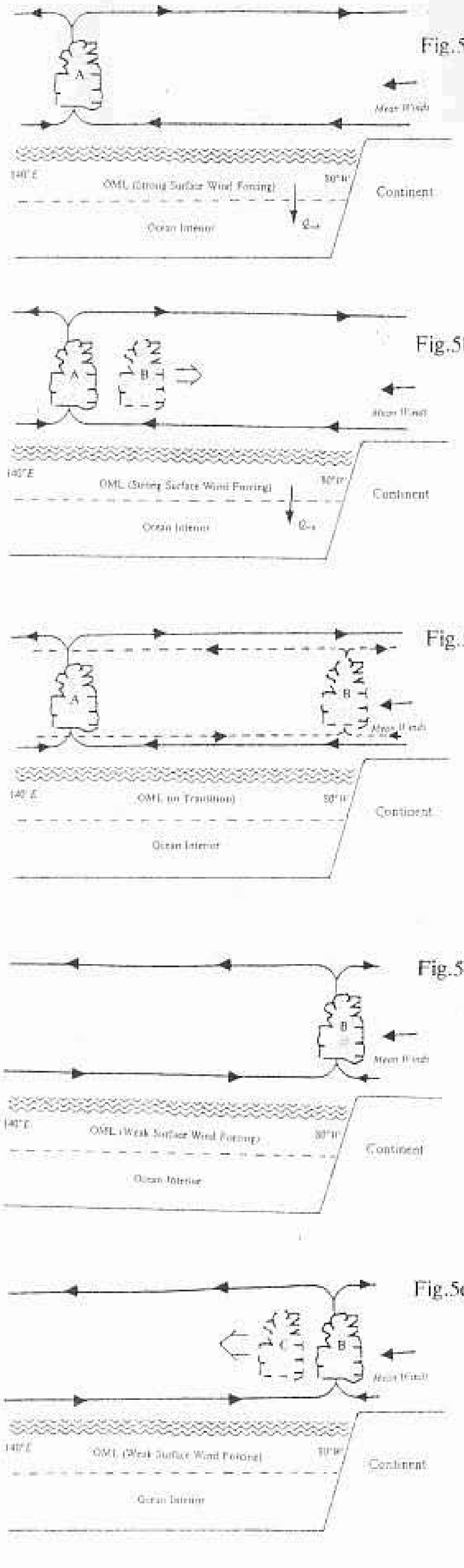
As Convection B continuously propagates (toward east), grows, and reaches the eastern Pacific, the La Nina enters its mature and transition stage (Fig.5c). During this stage, the strong surface wind stress is over the continent. Convection A and Convection B competes each other by reducing each other's low-level convergence and high-level divergence. The possibility of disturbance generation between Convection A and Convection B is greatly reduced due to this competition. Because Convection B is flowing over a relatively cool ocean surface (La Nina) with not too large growth rate (~ 1.8/*year*), two outcomes are expected: (1) Convection A survives, Convection B weakens and disappear; the system goes back to La Nina (Fig.5a). (2) Convection B survives, Convection A weakens and disappear; El Nino will take place (Fig.5d). This uncertainty makes La Nina cycle very irregular.

(c) Onset of El Nino

As Convection B survives (Fig.5d), the surface westerlies are prevailing over the tropical Pacific. They counterbalance the surface mean winds (easterlies) and make the total surface winds very weak. This will shift the OML to another regime (i.e., shallowing regime). The unstable convective disturbances (denoted by "Convection C") are generated in the west of Convection B and propagating westward (Fig.5e) with phase speeds ~ 0.4 m/s, and maximum growth rate (~ 3.5/year) appearing at the lowest wavenumber. Convection C is a fast growing mode.

(d) Mature of El Nino

As Convection C continuously moves toward west, grows, and reaches the eastern Pacific, the El Nino enters its mature stage (Fig.5f). During this stage, Convection C becomes very strong, and Convection B becomes very weak and disappears (Fig.5a) due to the high growth rate of Convection C and the warm SST. The certainty of disappearance of Convection B makes El Nino cycle relatively regular. The strong surface easterlies (from both the mean winds and Convection C) switch the OML back to the strong surface wind forcing regime; and La Nina starts again.



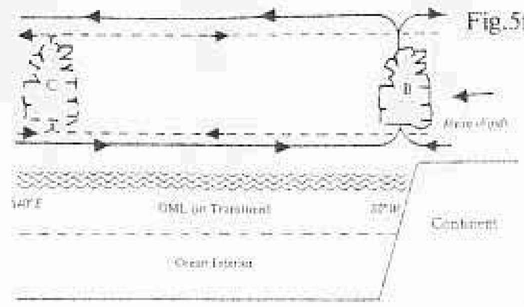


Fig. 5. El Nino and La Nina cycle.

7. SUMMARY

This study shows that a realistic thermodynamics for both atmosphere and ocean is needed in the ENSO theories. As the ocean surface is under weak wind forcing, the time rate change of SST perturbation is negatively correlated to the perturbation of the ocean mixed layer (OML) depth, $-h'_w$. However, as the ocean surface is under strong wind forcing, the time rate change of SST perturbation is positively correlated to the perturbation of the OML depth, h'_w. Such a difference leads to the generation of two different low frequency (interannual) modes propagating eastward (strong surface wind forcing) or westward (weak surface wind forcing). A new hypothetical theory about La Nina / El Nino cycle is presented.

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